Quaternary geology of the Duck Hawk Bluffs, southwest Banks Island, Arctic Canada: a re-investigation of a critical terrestrial type locality for glacial and interglacial events bordering the Arctic Ocean

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Abstract

Duck Hawk Bluffs, southwest Banks Island, is a primary section (8 km long and 60 m high) in the western Canadian Arctic Archipelago exposing a long record of Quaternary sedimentation adjacent to the Arctic Ocean. A reinvestigation of Duck Hawk Bluffs demonstrates that it is a previously unrecognised thrust-block moraine emplaced from the northeast by Laurentide ice. Previous stratigraphic models of Duck Hawk Bluffs reported a basal unit of preglacial fluvial sand and gravel (Beaufort Fm, forested Arctic), overlain by a succession of three glaciations and at least two interglacials. Our observations dismiss the occurrence of preglacial sediments and amalgamate the entire record into three glacigenic intervals and one prominent interglacial. The first glacigenic sedimentation is recorded by an ice-contact sandur containing redeposited allochthonous organics previously assigned to the Beaufort Fm. This is overlain by fine-grained sediments with ice wedge pseudomorphs and well-preserved bryophyte assemblages corresponding to an interglacial environment similar to modern. The second glacigenic interval is recorded by ice-proximal mass flows and marine rhythmites that were glacitectonized when Laurentide ice overrode the site from Amundsen Gulf to the south. Sediments of this interval have been reported to be magnetically reversed (> 780 ka). The third interval of glacigenic sedimentation includes glacifluvial sand and gravel recording the arrival of Laurentide ice that overrode the site from the northeast (island interior) depositing a glacitectonite and constructing the thrust block moraine that comprises Duck Hawk Bluffs. Sediments of this interval have been reported to be magnetically normal (< 780 ka). The glacitectonite contains a highly deformed melange of pre-existing sediments that were previously assigned to several formally named, marine and interglacial deposits resting in an undeformed sequence. In contrast, the tectonism associated with the thrust block moraine imparted pervasive deformation throughout all underlying units, highlighted by a previously unrecognised raft of Cretaceous bedrock. During this advance, Laurentide ice from the interior of Banks Island coalesced with an ice stream in Amundsen Gulf, depositing the interlobate Sachs Moraine that contains shells as young as ~24 cal ka BP (Late Wisconsinan). During deglaciation, meltwater emanating from the formerly coalescent Laurentide Ice Sheet deposited outwash that extended to deglacial marine limit (11 m asl) along the west coast of Banks Island. Our new stratigraphic synthesis fundamentally revises and simplifies the record of past Quaternary environments preserved on southwest Banks Island, which serves as a key terrestrial archive for palaeoenvironmental change.

Key words: Duck Hawk Bluffs, Banks Island, glacitectonics, Quaternary stratigraphy, Canadian Arctic glaciations and interglacials, paleoenvironments

Introduction

Previous reconstructions of the Neogene and Quaternary history of Banks Island, NT, have featured a complex and apparently continuous multiple glaciation record, notably from Duck Hawk Bluffs (DHB, Figs. 1a, b). The Banks Island stratigraphy purportedly includes late Neogene fluvial sand and gravel (assigned to the Beaufort Fm), overlain by the preglacial Worth Point Formation and then the deposits of at least three glacial and interglacial intervals (Vincent 1982, 1983, 1984, 1990; Vincent et al. 1983, 1984; Barendregt & Vincent 1990; Barendregt et al. 1998). This model was initially proposed for the surficial record of the entire island (70,000 km^2) where multiple till sheets and moraine systems, glacioisostatically controlled raised marine deposits and expansive proglacial lake sediments were assigned to three discrete glaciations, spanning at least the last 780 ka (Fig. 1b, c). Subsequently, expansive coastal sections were proposed to replicate the same stratigraphic record of the multiple glacial and interglacial sequences to the surficial geology (Vincent 1982, 1983). Fieldwork conducted during the past decade has proposed fundamental revisions of the surficial geology throughout Banks Island (England et al. 2009; Lakeman & England, 2012; Lakeman & England, 2013; Vaughan et al. this volume). In contrast to previously proposed models (Vincent 1982, 1983, 1984, 1990; Vincent et al. 1983, 1984; Barendregt & Vincent 1990; Barendregt et al. 1998), the revised surficial record of glacial and marine landforms were assigned to the Late Wisconsin. This has raised significant questions about the complexity and timescale of the previously reported stratigraphic record, given that Banks Island has been widely regarded as a critical type locality for glacial and interglacial events in the circumpolar Arctic (cf. Vincent et al. 1983, 1984; Clark et al. 1984; Matthews et al. 1986; Vincent 1990; Matthews & Ovenden 1990; Harrison et al. 1999; Barendregt & Duk-Rodkin 2011; Duk-Rodkin & Barendregt 2011; Li et al. 2011; O'Regan et al. 2011; Batchelor et al. 2012, 2013). Therefore it is timely that the stratigraphic record of Banks Island is reinvestigated, especially the primary sections at DHB and nearby Worth Point (Vaughan et al. this volume).

Study area and previous research

DHB is a continuous coastal cliff 8 km long and up to 60 m high, extending westward from Mary Sachs Creek to the southwest tip of Banks Island (Fig. 2). The bluffs compose an area of high land separating Amundsen Gulf to the south from Kellett River to the north. Small north-south orientated valleys dissect DHB, dividing it into five sectors designated: “Westernmost”, “West”, “Central”, “East” and “Easternmost” cliffs (Fig. 3). Cliff exposures to the east of Mary Sachs Creek were also investigated as they mark the western end of a moraine and till sheet (Sachs Till) deposited by the Laurentide Ice Sheet occupying Amundsen Gulf (Fig. 2, 3). Previous research assigned the moraine to the Early Wisconsinan, whereas landforms and sediments west of Mary Sachs Creek were assigned to the Bernard Till of the Banks Glaciation (>780 ka BP) overlain by a sequence of undeformed marine and interglacial sediments (Vincent 1982, 1983; Vincent et al., 1983, 1984).

The first lithostratigraphy presented for DHB was compiled by Vincent et al. (1983, Fig. 4). Their logs (A-I), correspond to our “Western” and “Central” cliffs, whereas east of DHB, their log J corresponds to our “Mary Sachs Creek cliff” (Fig. 3). Vincent’s model recognized seven major stratigraphic units within logs A-I, which were assigned formation status (Vincent et al. 1983). A prominent basal sand and gravel unit was originally assigned to the Neogene Beaufort Fm (Tb) based on its associated macroflora (Hills et al. 1974, Vincent et al. 1983). However, macrofloral differences between the Beaufort Fm type locality on Prince Patrick Island and the basal gravel at DHB (which appeared “more altered”), prompted Fyles (1990, p. 400) to designate the basal gravel at DHB as the “Mary Sachs gravel”. Overlying this gravel, Vincent et al. (1983) recognised the Worth Point Fm – a non-glacial, pre-Quaternary, aeolian, fluvial and lacustrine sand (unit 1, Fig. 4). This assumed a correlation with the type locality of the Worth Point Fm, ~ 30 km to the north that broadly occupied a similar stratigraphic setting (cf. Vincent 1980, 1982, 1983; Barendregt et al. 1998). Vincent further proposed that the Worth Point Fm at DHB was overlain by the Duck Hawk Bluffs Fm, marking the onset of glaciation comprised of the Bernard Till (Banks Glaciation) sandwiched between glacimarine sediments of the “Pre-Banks” and “Post Banks” seas (units 2a, b, c; Fig.

4). The Duck Hawk Bluffs Fm was reported to be capped by the Morgan Bluffs Fm (interglacial, unit 3), Nelson River Fm (full glacial ‘Big Sea’, unit 4) and Cape Collinson Fm (Sangamonian Interglacial, unit 5, Fig. 4). Despite the fact that these Fms at DHB are thinly bedded and discontinuous, with localised pockets of organics, they are nonetheless correlated with inferred interglacial and glacial deposits at Nelson River and Morgan Bluffs > 140 km to the east (cf. Vincent 1982, 1983; Barendregt et al. 1998). According to this model (Vincent et al. 1983), the most recent deposits (unit 6) are comprised of the “Pre-Amundsen Sea” glacimarine sediments and the “Sachs Till” that are assigned to the Early Wisconsinan Prince of Wales Fm. These deposits, however, only appear in Log J, overlying the Nelson River and Cape Collinson Fms (Fig. 4).

Compilations of organic samples and sediments used for dating and paleoecological reconstructions in the original model for DHB can be found in Vincent et al. (1983, 1984), Matthews et al. (1986), Vincent (1990), Barendregt and Vincent (1990) and Barendregt et al. (1998; Fig. 4). The stratigraphic logs and paleomagnetic measurements were complied into a composite lithostratigraphy (Barendregt and Vincent, 1990; Fig. 1c). Paleomagnetic measurements indicate that sediments in the Worth Point Fm, Duck Hawk Bluffs Fm, and lower Morgan Bluffs Fm are magnetically reversed (> 780 ka, Matuyama Chron), whereas all sediments from the upper Morgan Bluffs Fm to the surface are magnetically normal (< 780 ka, Bruhnes Chron; Vincent et al. 1984; Barendregt et al. 1990).

Methods

Stratigraphic exposures at DHB were documented using annotated photomosaics and vertical profile logs, which included primary sedimentary structures, bed contacts, sediment body geometry, sorting, texture and organic macrofossils. These data were then employed in the characterisation of individual lithofacies, classified according to the facies codes proposed by Eyles et al. (1983) and Evans and Benn (2004). Where relevant, secondary sedimentary structures, including faults, folds, ice wedge pseudomorphs and cross-cutting intrusions or clastic dykes, were also entered onto stratigraphic logs and photomosaics. The orientations of the dipping surfaces of thrust fault planes were also measured and entered onto vertical profile logs as individual data points or as great circles on lower hemispheric stereoplots wherever numerous measurements were possible.

Former debris transport pathways were evaluated through clast form analyses on predominantly sandstone, quartize and chert lithologies, which included Powers roundness (VA = very angular; A = angular; SA = sub-angular; SR = sub-rounded; R = rounded; WR = well rounded) and clast shape (see Benn, 2004). Roundness was assessed visually using histogram plots and statistically by calculating an RA value (relative angularity = percentage of clasts in the VA and A categories), an RWR value (percentage of clasts in the R and WR categories; Benn et al., 2004; Lukas et al., 2013) and an average roundness value, wherein VA=0, A=1, SA=2, SR=3, R= 4 and WR=5 (cf. Spedding and Evans, 2002; Evans, 2010). Clast shape was analysed statistically by using clast shape triangles (Benn, 2004) from which C40 indices (percentage of clasts with C/A axial ratios ≤0.4) were derived and compared to RA, RWR and average roundness values in co-variance plots following procedures outlined in Benn and Ballantyne (1994). The RWR index is employed here because previous studies have reported that glacifluvial reworking of clast forms results in the failure of the RA-index to discriminate between different transport pathways (Benn et al. 2004; Lukas et al. 2013, in press; Evans et al. 2010).

Control samples for clast form assessment are normally collected from material derived from known processes operating in the vicinity and using similar lithologies to those sampled in stratigraphic section. However, DHB lacks a glacierized catchment (glacifluvial, subglacial and slope processes); requiring that clast form analysis be taken from existing databases, using lithologies similar to those sampled locally.

We employ the "Type 2" co-variance plot of Lukas et al. (2013), which represents mostly highly anisotropic lithologies (Fig. 5).

Clast macro-fabrics were measured on diamictons using predominantly the A/B planes and in some cases also the A-axes of clasts (n=50 or 30), which are thought to rotate towards parallelism with the principal axis of extensional strain in a deforming medium or with the plane of slip during brittle deformation (Benn and Evans, 1996). The data were processed in Rockworks stereonet software and depicted using Schmidt equal-area lower hemisphere projections based of spherical Gaussian distributions. The macrofabrics were then analysed for strength, modality and isotropy following procedures outlined by Benn (1994, 2004), thereby facilitating an assessment not only of the direction of applied stress but also the genesis of the deposit. The latter was determined through comparisons with the clast macrofabrics of subglacial tills sampled at modern glacier margins, subaqueous glacigenic diamictons and glacitectonites, utilizing the data presented by Benn (1994, 1995), Evans and Hiemstra (2005) and Evans et al. (2007) and employing specifically the modality/isotropy plot of Evans et al. (2007; Fig. 6).

Organics

Twelve organic samples were collected and analysed from the DBH sections. Bulk organics were subsampled then submerged in water, rinsed and screened through two sieves (710µm and 450µm) for macrofossils. From the retained residue, preserved subfossils were examined with dissecting (Wild M5A) and compound (Leitz Laborlux S) microscopes. Bryophytes were determined using Nyholm (1956 -1965) and Lawton (1971) with nomenclatural adjustments using Crosby et al. (1999). Species determinations of the Calliergonaceae and Amblystegiaceae used the revised interpretations by Hedenäs (1993, 2006). Paleoecological reconstruction utilises Kuc and Hills (1971), Kuc (1974), Steere (1978), Steere and Scotter (1979), and Janssens (1983). Macrofossil vouchers were mounted on permanent slides and deposited in the Cryptogamic Herbarium (ALTA), Biological Sciences, University of Alberta.

Results

Sedimentology and stratigraphy

The sedimentological and stratigraphic data for DHB are compiled in vertical profile logs and annotated photographs (Fig. 7). The descriptions and interpretations of the lithofacies associations (LFA) recognised throughout the bluffs are then presented, employing details illustrated in Figure 7 and referring to the clast form and macrofabric data analysis (Figs. 5, 6).

i) Description

LFA 1 comprises up to 38 m of predominantly well-sorted, cross-stratified sand and gravel arranged in stacked sequences of horizontal and planar bedding structures. The most common lithofacies are planar and trough cross-bedded sand, often with granule lags, and horizontal to planar or trough cross-bedded gravel (Fig. 8). Also common, especially towards the base of the exposures, is matrix-supported gravel with boulders. Macrofossils, commonly comprising crudely bedded organic detritus, also include large tree debris (i.e., logs or stumps with degraded root balls) and compressed mats of small woody fragments, occur throughout the sequence (Fig. 8a). The largest concentrations of tree fragments and stumps occur in coarser grained gravelly beds, particularly in matrix-supported gravel with sand, silt and clay intraclasts. The floral and faunal assemblages within LFA 1 are diverse, as illustrated by lists reported in Roy and Hills (1972), Hills et al. (1974), Hills (1975), Matthews et al (1986) and Matthews (1987).

Sandier lithofacies occur towards the top of LFA 1, where they record a fining upward sequence. In the central cliffs and West Log D these upper beds contain rhythmically bedded (varve-like) silt and clay with lonestones interbedded with matrix-supported gravelly scour fills containing a distinct horizon of logs and woody detritus. At Central Log A, the underlying LFA 1 gravel is arranged in prominent clinoforms or foreset beds dipping at 20° towards the southwest (Fig. 8c). Well-developed ice wedge pseudomorphs, up to 5 m deep, also occur towards the top of LFA 1 in some sections (Fig. 7a iii). Clastic dykes or fissure fills are also common, characterised by crudely, sub-vertically bedded and poorly sorted sand and gravel that in some cases is clearly rooted in underlying strata. Some clastic dykes are also characterised by branching and upward extending limbs or tenticles that wedge out in the host materials.

Palaeocurrents measured on sand and gravel bedforms throughout LFA 1 indicate progradation predominantly towards the west, with the most common flow directions towards the WSW and WNW. Clast lithologies are dominated by chert, quartzite and sandstone with minor amounts of shale towards the top of LFA 1; quartzite dominates the counts in the middle of LFA 1. Striae are visible on a small number of clasts, ranging from 16% at the base to 6% in the middle and upper lithofacies. Clast form data from LFA 1 reveal a vertical increase in rounding (2.28 to 3.12) and RWR values (2-28%), reasonably consistent C40 values throughout (26-50%), and a vertical reduction in, but generally low, RA values (10-0%). Exposures through LFA 1 in the eastern cliffs reveal large scale reverse (thrust) faults that displace underlying Kanguk Fm and overlying LFA 1 beds towards the southwest (Fig. 7d iv-viii).

LFA 2 is only locally preserved and often heavily deformed by simple shear structures (e.g. Fig. 7c iii, inset photos 5-9 & Fig. 7d iii, inset photos a-k). It comprises 11 m, but generally less than 8 m, of rhythmic, planar to horizontally bedded sand with thin and discontinuous beds of massive granule gravel or lags and laminated fine sand, silt and clay arranged in stacked units and interbeds. Outsized clasts or lonestones occur in many finer grained units. At the base of LFA 2 (west cliffs) a laterally extensive surface (~ 100 m wide) composed of locally disharmonically folded laminae also contains prominent ice wedges and thick peat composed of *in situ* bryophyte assemblages characterised by excellent preservation (Fig. 9). Nine bryophtye dominated samples were collected, five of them spanning a single 1.5 m section of interbedded organics with fine sands generally accordant with the ice wedge surface.

Secondary structures include clastic dykes intruded upwards from underlying LFA 1 gravel, sand wedge pseudomorphs (Fig. 9), intraformational burst-out clastic dykes and zones of overfolded or crumpled bedding and thrust faults associated with boudinage or high fissility/pseudo-lamination (Fig. 10). Deformation and intense modification and/or erosion are locally the most prominent sedimentary signatures, for example at the junction of LFA 1 and LFA 3 near west Log D, where LFA 2 has been pinched out from an initial thickness of approximately 6 m (Fig. 11). At this location, LFA 2 displays a 1-2m thick basal zone comprising highly attenuated inter-digitated beds of gravelly to clay-rich diamictons and sand/silt/clay rhythmites. This passes upwards into rhythmites containing diamictic intraclasts with sharp and angular boundaries; significantly the rhythmite bedding drapes the intraclasts and in places is deformed beneath them. The sense of shearing in the deformation structures (Figs. 7c iii, 10) is predominantly towards the southwest, although some structures indicate shearing towards the north, specifically ranging from between NNW and NNE. The southwesterly shearing direction is recorded in thrust faults that continue into, or are developed within, underlying LFA 1 and overlying LFA 3.

LFA 3 varies in thickness from 6 to 14 m and is separated from underlying LFA 2 by a heavily deformed zone characterised by intensely folded, thrust faulted and/or highly fissile sand, silt, clay laminae or an erosional contact associated with attenuated rafts (intrabeds) of LFA 2 sediments (e.g. Fig. 7c iii, inset photo 5; 7b iv, inset photo 2). It comprises a massive to pseudo-laminated, matrix-supported diamicton (Fig. 7d i, inset photo 1) with localised zones of discontinuous, stratified sand and/or gravel lenses up to 1 m thick (Fig. 7c iii, inset photo 4). Areas of more densely spaced stratified lenses constitute an interbedded relationship with the diamicton. Zones of attenuated and/or overfolded lenses indicate post-depositional deformation of the diamicton and tend to be concentrated at the top (Fig. 7b iv, inset photo 3) or base of LFA 3 (Fig. 7d i, inset photo 2). The sense of shearing in the lower part of LFA 3 is variable but predominantly towards the southwest, with subsidiary thrusts towards the north (between NNW and NNE). In the upper part of LFA 3 the sense of shearing is towards WSW or west, although there is also some evidence of a northerly shear direction. Clast macrofabrics are moderately to well clustered ($S1 = 0.451 – 0.773$) and display a range of orientations, the most prominent stress towards the southwest but also including N-S alignments. Unusually steep A/B plane dips occur in the recumbent folds of lower LFA 3 in west cliff Log D (WCD F1), although this sample is aligned NW-SE. Clast macrofabric strengths, as quantified by the fabric shape ternary plot (Fig. 6a) and the modality/isotropy plot (Fig. 6b), are variable, with the strongest clusters being represented by the A axis and A/B plane data from the upper part of LFA 3 (WCD F3). In the ternary graph, the macrofabric shapes plot across the spectrums of the glacitectonite and subglacial traction till envelopes (Fig. 6a). Clast lithologies vary depending on sample location. A more restricted lithological component (chert, quartzite and sandstone) occurs in the deformed contact zone with underlying LFA 1 and in areas characterised by attenuated gravel lenses. A more varied clast lithology, comprising chert and quartz with minor components of sandstone, shale, granite, gabbro and limestone, occurs towards the middle and in the more diamictic zones (LFA 3).

LFA 4 comprises less than 5 m of sand, silt and clay arranged in horizontal cross-laminae, rhythmites, climbing ripples or draped laminae that locally contain organic detritus. The upper and lower contacts of LFA 4 are commonly characterised by heavily brecciated clay or contain boudinage and thrust faulted laminae (Fig. 12, panels a & c). Some outcrops of LFA 4 also display widespread deformation in the form of overfolded bedding and thrust faults, although normal faulting and open folding or convolute bedding structures are also evident (Fig. 7c iii, inset photo 3). In some sections, LFA 4 comprises only a thin (< 2 m thick) bed of brecciated clay (Fig. 7b iv, inset photo 4) or laminated sand displaying overfolds and sheath folds. The sense of shearing, as recorded by thrust faults, is generally towards the SSW but southerly and southwesterly orientations are also evident. Numerous examples of narrow, anabranched clastic dykes rise sub-vertically through the thickest outcrop of the sandy, climbing ripple lithofacies (LFA 4, West Log A) and have created offset beds between the blocks within the host material (Fig. 12, panel b). The contact of LFA 4 with LFA 3 (West Log A) is marked by a clast lag. Although the association is predominantly heavily deformed, the most unaltered and thickest outcrop in West Log A displays climbing ripple drift indicative of a palaeocurrent from the south or southwest (Fig. 7b ii).

LFA 5 comprises 5-10 m of tabular sets of horizontally bedded to massive gravel, interbedded with horizontally bedded or cross-laminated sand and occasional units of matrix-supported gravel (Fig. 13a, b). The thickest outcrop displays a general coarsening-upward sequence from well-sorted, horizontally bedded granule gravel and sand to matrix-supported and less well sorted, horizontally bedded cobble to boulder gravel. Clast lithologies are dominated by quartzite but contain subsidiary amounts of chert and sandstone with minor limestone. LFA 5 differs from LFA 1 based on its general coarsening-upward characteristics, coarser grain size and the presence of limestone clasts. Clast form data from LFA 5 indicate a sub-rounded sample (average roundness = 2.22 & 2.46; RA = 8 & 24%; RWR = 2 & 6%), with predominantly blocky shapes (C_{40} = 30 & 42%) and striae visible (< 14% of clasts). Upper and lower contacts are sharp but are locally interdigitated or amalgamated with LFA 6 (Fig. 7b iv, inset photo 5) and LFA 4 (Fig. 13d), respectively. A heavily deformed, discontinuous bed of silt/clay rhythmites occurs between LFA 5 and LFA 6 in West Log D, the base of which has been amalgamated with the gravel at the top of LFA 5 (Fig. 13c). Gravel in Easternmost Cliff has an unclear relationship to the primary sections at DHB to the west. This gravel, designated LFA 5a, rests uncomformably on LFA 1 and is not capped by LFA 6 as elsewhere (Fig. 7e). Furthermore, there is no evidence of deformation and palaeocurrents are oriented towards the NW.

LFA 6 crops out along the length of DHB and varies widely in thickness (1 to 11 m), displaying the most complex deformation structures of any of the sediment bodies at DHB. It comprises a predominantly heterogeneous diamicton or a mélange of discontinuous interbeds of massive, matrix-supported diamicton, matrix-supported gravel, and laminated sand, silt and clay with localised attenuated lenses of degraded organics (Fig. 7a iv-vi, 7b i, iv, 7c ii, iii). Three samples of highly compressed and fragmented detrital material displaying poor preservation were collected for identification. Collectively, LFA 6 has the appearance of a stratified diamicton but locally it displays a more massive to pseudo-laminated character, where sand, silt and gravel lenses and occasional gravel lags occur as thin beds, stringers or wisps. Some laminated sand, silt and clay bodies appear to be rafts within the diamicton (intraclasts) due to their heavily deformed internal bedding and deformed contacts or attenuated appearance. Basal contacts with other lithofacies associations are typically erosional or deformed, and associated with fissility, boudinage and pseudo-lamination in sediments on either side of the contact. Basal diamictons can also display a higher concentration of gravel clasts that locally result in a clast-supported character or even a several metre-thick, highly contorted mélange of diamicton and stratified sediment bodies. In some sections the whole of LFA 6 has been complexly deformed to produce stacked overfolds and thrust faults indicative of shearing towards the southwest or WSW. Clast macrofabrics are only moderately clustered (S_1 = 0.456-0.637) but indicate stress directed predominantly towards the SW. Clast macrofabric strengths, as quantified by the fabric shape ternary plot (Fig. 6a) and the modality/isotropy plot (Fig. 6b), are variable and range from bi-modal to multi-modal. Overall the macrofabric shapes in the ternary graph plot across the spectrums of the glacitectonite and subglacial traction till envelopes but there is a trend of changing fabric strength vertically through LFA 6. The strongest clusters are from the basal diamictic zones (e.g. CCA F1 & F3, A/B plane data) whereas the weakest are from the top of the association (e.g. WCA F3 A-axes & CCA F2 A/B planes) where relatively low strain deformation indicators (open and recumbent folds) are ubiquitous.

Clast lithologies in LFA 6 are dominated by quartz, chert and sandstone, although one diamict reveals a concentration of sandstone lithologies at the expense of chert. Clast forms similarly reflect the nature of the sampled materials, whereby the gravelly sample (CCA 2) was characterized by relatively low C40 (38%) and RA (12%) values and a high average roundness (2.56) compared to the same values for the diamicticts of LFA 6 (C40 = 50-74%; RA = 38-52%; average roundness = 1.74-2.04). A small number of clasts (<10%) in LFA 6 were striated.

ii) Interpretation

LFA 1. The thick, tabular sequence of well sorted, cross-stratified sand and gravel records deposition by a former glacifluvial braided river typified by fluctuating discharge (e.g. Church 1974; Boothroyd & Ashley 1975; Miall 1978, 1985, 1992; Marren 2005). The highest discharge is recorded by matrix-supported gravel with boulders indicative of hyperconcentrated flows or a traction carpet (Maizels 1989a, b; Siegenthaler & Huggenberger 1993; Mulder & Alexander 2001). These are separated by, and vertically give way to, lower discharges recorded by planar or trough cross-bedded sand and horizontal to planar or trough cross-bedded gravel. These deposits record aggradation and migration of gravel bars and sheets and downstream accretionary macroforms with relatively tightly constrained palaeocurrent indicators (Miall 1977, 1985; Collinson 1996). The general lack of scour fills and an overall fining-upward is indicative of an aggrading system initially characterized by high magnitude/high frequency events that were gradually replaced by low magnitude/high frequency events (Marren 2005). Partially to heavily-scoured remnants of laminated and massive fines represent deposition during low flow conditions in abandoned shallow channels. The palaeocurrent data from LFA 1 indicate that the outwash streams were flowing westerly, away from a glacier source located in Amundsen Gulf. This glacier likely eroded the abundant LFA 1 lithologies (chert, quartzite and sandstone) from Mesoproterozoic and Neoproterozoic bedrock widely distributed to the east and south (western Victoria Island, Amundsen

Gulf and the adjacent mainland; Harrison et al. 2013). These lithologies are consistent with the advance of the NW Laurentide Ice Sheet through Amundsen Gulf (cf. Dyke et al. 2002; Batchelor et al. 2012; MacLean et al., 2013).

The crudely bedded organic detritus and degraded tree fragments (stumps) in LFA 1 are clearly reworked (cf. Matthews et al. 1986; Fyles 1990; Vincent 1990). The finer grained detritus was deposited in concentrated pockets during waning flow stages, whereas horizons of logs and larger tree fragments were emplaced with the coarser grained and poorly sorted sediments associated with hyperconcentrated flows or traction carpets during high discharge. These reworked organics also differ significantly from the type location of the Beaufort Fm., a characteristic that prompted Fyles (1990) to distinguish them as ‘the Mary Sachs Gravel’. Notably, the enclosing gravel and sand of LFA 1 is much coarser and exhibits an entirely different fluvial architecture than the sandier Beaufort Fm. The composition and degraded nature of the LFA 1 organics suggests that they were eroded by glaciers or their meltwater from Neogene and/or Paleogene sediments presumably to the southeast of Banks Island (e.g., Fyles, 1990; Fyles et al., 1994). The proximity of a glacier is highlighted by the preservation of striae on individual clasts, especially in the very coarse, lower gravel. The vertical decrease in striae preservation and angularity (RA), and concomitant increase in roundness and upward-fining in LFA 1, is indicative of an increasingly distal ice margin. The co-variance plots (Fig. 5) indicate a strong fluvial signature, especially in the C40/RA data, but subglacial characteristics are also evident, particularly in the C40/RWR data for the coarser grained and more matrix-supported gravel, as would be expected in ice-proximal settings.

It is important to note that the widespread thrusting of LFA 1 and underlying Kanguk Formation towards the southwest occurred post-depositionally. This glacitectonic thrusting was produced by Laurentide ice that advanced from the Kellett River valley (interior Banks Island), during the emplacement of LFA 6 (see below). The injection of clastic dykes and fissure fills during this glacitectonic disturbance was due to the overpressurizing of groundwater in gravel aquifers, likely in taliks. The branching limbs of many of the dykes are indicative of burst-out structures diagnostic of hydrofracturing by vertically escaping groundwater pressurized by glacier overriding (cf. Rijsdijk et al. 1999; LeHeron & Etienne 2005).

LFA 2. At the top of LFA 1, rhythmites with lonestones, overlie prominent clinoforms or foreset beds, recording a change from subaerial glacifluvial to subaqueous deposition. This was initially on a shallow delta front, prograding foreset beds in a southwesterly direction followed by suspension sedimentation with dropstones, presumably from floating ice in a marine or lacustrine basin. The occurrence of matrix-supported gravelly scour fills with logs and woody detritus suggests pulses of cohesionless sediment gravity flows into the basin. Overlying the rhythmically-bedded sand, prominent ice wedge pseudomorphs (up to 4 m deep; Fig. 9a-c; cf. Mackay & Burn 2000) are interspersed with thick (>1 m) sections of compressed *in situ* bryophytes (Figs. 9d, 7e) indicating the establishment of a tundra surface. Four bryophyte families and 10 distinct species have been determined from LFA2, with the most abundant material from the Calliergonaceae and Campyliaceae (Table 1). These macrofossils are exceptionally well preserved, including material with intact, fragile alar cells typical of these two families (Fig. 14c-f). The data from these autochthonous deposits indicate a wetland environment, characterized as a rich fen. This ecosystem and habitats therein are consistent with species found in the extant flora of Banks Island (Fig. 14b). The autecology from this range of species suggest hydric to mesic habitats including standing pools, wet meadows, and microhabitats created by ice wedges. Despite the predominance of bryophytes in LFA 2, these macrofossils had not been previously documented at DHB. In contrast, a minimal proportion of the samples (< 5%) consisted of vascular plant material (including Cyperaceae seeds, gramminoid leaves and woody twig fragments). Collectively, the ice wedge pseudomorphs and vegetation within LFA 2 represent an isochronous terrestrial surface of interglacial rank.

Following the formation of ice wedges, the injection of intraformational clastic dykes and fissure fills in LFA 2 relates to the overpressurizing of groundwater in gravel aquifers or taliks due to glacier overriding. The branching dyke limbs that emanate from underlying LFA 1 gravels and sands are indicative of the burst-out structures that develop during hydrofracturing (cf. Rijsdijk et al. 1999; LeHeron & Etienne 2005). A variant of the gravel and sand filled clastic dykes occurs within the sequence of glacitectonite and rhythmites near west Log D (Fig. 11). Here the vertically-widening fracture is characterized by a mosaic of partially disaggregated blocks of the host materials in its widest zone where it contains gravels injected from underlying LFA 1. Because these gravels only occur in the basal narrow neck, they presumably record the rapid reduction in flow competence during hydrofracture filling.

Deformation structures, including overfolded or crumpled bedding and thrust faults associated with boudinage or high fissility/pseudo-lamination, are representative of simple shear and were also induced by glacitectonic disturbance (cf. van der Wateren 1995; McCarroll & Rijsdijk 2003). This predominantly records a glacier advance from the northeast. Subsidiary shear indicators record deformation from the south or southeast; but these occur only at the top of LFA 2 and the base of LFA 3, suggesting earlier emplacement (see below). The development of zones of boudinage, shallow thrust faults, overfolding or bed crumpling at different levels within LFA 2 - especially at the contact separating sediments of different grain size - represents deformation partitioning due to contrasting rheological properties. Blind thrusts were also initiated in underlying LFA 1, increasingly towards the central and eastern cliffs where large-scale deformation is more prevalent (Fig. 7c, d). A zone of intense deformation is recorded by a lens of LFA 3 diamicton in fault contact with LFA 2 (east cliff Log A, Fig. 7d i). This constitutes a type B glacitectonite (non-penetratively deformed, pre-existing sediments; *sensu* Benn & Evans 1996) developed during the emplacement of LFA 3 (see below). A similar glacitectonite occurs at the junction of LFA 1 and LFA 3 near west Log D (Fig. 11). Bed attenuation by deformation is best developed in lower LFA 2 where the greater differences in sediment grain size, and hence rheology, exist.

LFA 3. Within LFA 3, locally abundant, stratified lenses indicate subaqueous deposition. These stratified sediments were subsequently thrust faulted, folded and attenuated - most evident in the upper and lower zones of LFA 3. This deformation was likely responsible for the intraclast boudinage and pseudo-lamination in the surrounding diamicton (e.g. Hart & Roberts 1994; Roberts & Hart 2005; Ó Cofaigh et al. 2011). This process was also associated with the attenuation of LFA 1 and 2 sediments within the heavily deformed base of LFA 3, some of which may have been cannibalized and rafted by the deformation process within a shear zone at the contact of LFA 3 with older sediments. The shearing direction from the south and southeast, recorded in lower LFA 3, also impacted upper LFA 2, whereas the more prominent shearing direction throughout LFA 3 is from the northeast (also recorded in LFA 4). The local development of a heavily deformed zone at the top of LFA 3 was likely imparted during the shearing of overlying LFA 4 sediments. Although clast macrofabric strengths from the LFA 3 diamicton are variable, they possess shapes that compare with glacitectonites and subglacial traction tills and, in some cases, a modality/isotropy signature that reflects high lodgement components.

The similar orientation of the macrofabrics and shear indicators, are considered diagnostic of a glacitectonite derived from stratified sediments (e.g., interbedded mass flow diamictons, gravelly mass flows and localized traction current sands and suspension deposits). This evidence indicates that LFA 3, like all the other LFA's and the underlying Kanguk Fm. bedrock, has been displaced by a glacier advancing from the NNE or NE (i.e., Laurentide ice crossing the interior of Banks Island). Minor south to north shearing indicates that LFA 3 and upper LFA 2 were also disturbed by glacier ice flowing northwards (i.e., Laurentide ice advancing through Amundsen Gulf). The appearance of granite and gabbro clasts within the LFA 3 diamict records the arrival of far-travelled Laurentide ice from mainland Canada. Clast form co-variance strongly indicates a subglacial (abraded) source for LFA 3 with a minor fluvial component (Fig. 5), including glacifluvial clasts locally cannibalized from LFA 1 by overriding ice.

LFA 4. LFA 4 has restricted outcrops, is heavily deformed and contains organic detritus that is clearly allochthonous and highly degraded. Nonetheless, a variety of indicators (fine-grained, horizontally bedded cross-laminae, rhythmites, climbing ripples and draped laminae) demonstrate that LFA 4 was deposited in a subaqueous environment with palaeocurrents from the south or southwest. Traction currents were likely responsible for the development of the clast lag at the contact between lower LFA 4 and underlying LFA 3 (west cliff, Log A; Fig. 7bii). LFA 4 coarsens upwards, indicative of basin infilling and/or an approaching sediment source. The narrow, anabranched clastic dykes in the sandy, climbing ripples (LFA 4, west cliff, Log A) record hydrofracturing by the upward injection of overpressurized groundwater during shearing. The branching limbs of the dykes are similar to the burst-out structures reported by Rijsdijk et al. (1999) but differ because they are composed solely of sand derived from the walls of the host sediment, rather than from an underlying gravel aquifer (cf. LeHeron & Etienne 2005). Hence the pressurized water was generated at the contact between the sand and underlying impermeable brecciated clays. The development of thin zones of brecciation, boudinage, sheath folding or thrust faulting at the upper and lower contacts of LFA 4 relates to the localized partitioning of deformation controlled by the location of clay/silt beds. The juxtaposition of this style of deformation with the larger scale overfolds and thrust faults, developed in the extensive bodies of sandy lithofacies, is indicative of a Type B glacitectonite (*sensu* Benn & Evans 1996). This was constructed by the overriding of glacier ice advancing from the north or northeast. These shear directions are compatible with those developed in the underlying LFA's.

LFA 5. The primary characteristics of LFA 5 are typical of gravel sheet and occasional traction-carpet sedimentation in a ice-proximal sandur (e.g. Boothroyd & Ashley 1975; Miall 1978, 1985, 1992; Marren 2005; Maizels 1989a, b; Siegenthaler & Huggenberger 1993; Mulder & Alexander 2001). Glacier proximity is supported by the subglacial to fluvial signature of the clast forms (Fig. 5). The coarsening-upward sequence, including striated clasts, culminates in the arrival of a glacier that overrode the site, depositing LFA 6. Post-depositional shearing within underlying LFA 4 is recorded by attenuation and interdigitation at the contact with LFA 5, imparted during deposition of LFA 6. The contact between LFA 5 and 6 reflects the development of a shear zone where upper LFA 5 sediments have been cannibalized, producing a mélange of crudely stratified diamictons with discontinuous interbeds of matrix-supported gravel, laminated sand, silt and clay (LFA 6, see below).

LFA5a. At Easternmost Cliff, glacifluvial gravel (LFA 5a) unconformably overlies LFA 1 (Fig. 7e) rendering uncertain its relationship with respect to LFA 5 in the main sections of DHB to the west. The fact that LFA 5a extends to the modern surface, appears undeformed and has paleocurrents oriented to the NW, requires that it is younger than LFA 5. Indeed, we interpret LFA 5a as deglacial outwash deposited from retreating Laurentide ice in the Kellett River valley and Amundsen Gulf at the end of the last glaciation (see Discussion).

LFA 6. Based on its heterogeneity, strongly orientated clast macrofabrics, and macroscale deformation structures, LFA 6 constitutes a Type B glacitectonite (*sensu* Benn & Evans 1996). However, zones of more highly strained Type A glacitectonite (*sensu* Benn & Evans 1996) at the basal contact with older LFA's are recorded by intraclast/interbed attenuation (pseudo-lamination), strong boudinage or fissility, and strengthened clast macrofabrics. The production of this mélange is attributed to two separate processes: 1) primary sedimentation as interbedded and interdigitated mass flow debris, fluvial gravel and sand, and subaqueous suspension sediments; and 2) cannibalization of underlying stratified sediments (LFAs 4 and 5). This produced a continuum of forms ranging from low strain fold structures to attenuated rafts to tectonic lamination (e.g., Fig. 7). The discontinuous lenses of degraded organics (black stems and woody twigs, gramminoid leaves, whitened rootlets) within LFA 6 are rafts and include seven species of bryophytes (Table 1). The material was highly compressed, and difficult to separate into distinct taxa. Indeed, one of the three samples analysed was too degraded to recover any taxa. Much of the material consisted of only fragmented leaves and only the minute taxa had stems with intact leaves. The bryophytes that characterise the LFA 6 samples represent hydric (*Calliergon* spp. *Tomenthynum nitens*) to mesic (*Ditrichum flexicaule*, *Dicranum* sp.) habitats. Three species collected from LFA 6 were also present in LFA 2, whereas four others were not found in LFA 2, but are common in the modern flora. The poor preservation quality suggests that the material has been redeposited from unknown sources, and is clearly allochthonous.

The predominant shearing direction within LFA 6 records an ice advance from the interior of Banks Island (NE). This has resulted in complex folding and thrusting and localized thickening of LFA 6. This shearing direction is recorded throughout underlying LFAs 1-4 as well as the Kanguk Formation, demonstrating deep-seated but partitioned glacitectonic disturbance imparted during the emplacement of LFA 6. This strain signature was therefore overprinted on the northerly/northeasterly aligned shearing direction recorded in LFA 3 and upper LFA 2 by the earlier, northerly flowing ice that deposited LFA 3. One anomalous clast macrofabric (northerly, CCA F1) occurs in the basal mélange of LFA 6. However, this sample is from an isolated outcrop (Fig. 7c i), rendering its apparent shearing direction equivocal.

Sedimentology and stratigraphy (Mary Sachs Creek cliff)

The cliffs east of Mary Sachs Creek (Figs. 2b, 3) provides a cross-section through the seaward margin of the coast-parallel Sachs Moraine ("section J", Vincent et al. 1983; Vaughan 2013). We provide a summary of two logs based on a reconnaissance survey of these sediments. Log A is closest to Mary Sachs Creek whereas Log B occurs ~ 1 km farther east (Fig. 3, 15a & b). The sediments exposed in Log A appear to descend eastward and disappear below the modern beach and therefore are assumed to stratigraphically underlie the sediments of Log B. A sand and gravel- bench caps both logs at ~ 15 m asl, incised into the seaward margin of the Sachs Moraine.

i) Description

Log A contains a sequence of upward-fining, highly contorted, cross-stratified sand and fine gravel to rhythmically bedded sand, silt, clay with minor pebble gravel and sporadic clots of gravelly diamicton (Fig. 15a). An ice wedge was initially developed in the lower cross-stratified sand and fine gravel before they were folded and thrust faulted. The deformed, ice wedge pseudomorph is unconformably overlain by a relatively undisturbed lens of laminated silty clay with dropstones, interbedded with gravelly diamicton, that contains a bed (10 cm) of sandy, black organics containing wood fragments (Fig. 15a). This organic material occurs repeatedly up-section as discontinuous stringers. Log A is capped by 1-2 m of poorly exposed and heavily cryoturbated, gravelly diamicton. The general sense of displacement on shallow reverse faults, overturned bedding and boudinage structures is from the southeast.

Log B (12 m) displays a coarsening-upward sequence of horizontally bedded, sandy granule gravel with occasional silty sand laminae (Fig. 15b iii) that grades to poorly-sorted cobble to boulder gravel and bouldery matrix-supported gravel (Fig. 15b ii). This is capped by laminated to crudely stratified, cobbly diamicton that is silt and clay-rich, abruptly and conformably overlain by interlaminated silt, clay and pebbly silty sand (Fig. 15b i). Although the bedding on these units dips towards the NW, individual laminae do not thicken in that direction, indicating post-depositional tilting. The uppermost part of the log (underlying the 15 m bench) contains ≤ 2 m of clay-rich, massive diamicton interfingered with the underlying laminated sediments and cobbly diamicton .

ii) Interpretation

Log A. Broadly, the sedimentology of log A, including the deformed ice wedge pseudomorph, is similar to LFA 2 in DHB (e.g., central cliff Log A, Fig. 7c iii). Prior to glacitectonic disturbance, the overlying organic bed (Fig. 15a) was redeposited into a sequence of subaqueous silt/clay rhythmites containing dropstones and gravelly mass flows. The highly attenuated shear margins bracketing the organic bed in

Log A demonstrate that it has been displaced as a tectonic raft. Previous research proposed that the organic beds - similar to those described in Log A - be assigned to the last interglacial (locally called the Cape Collinson; Vincent 1982; 1983). Because these organic beds are clearly a raft, they cannot constitute an iscochronous surface, precluding chronostratigraphic significance (see Discussion). Shearing of the Log A sediments from the southeast documents an ice advance from Amundsen Gulf, corresponding to the earliest of two primary shearing directions recorded at DHB (imparted during the emplacement of LFA 3). Therefore, these deformed deposits are interpreted as a Type B glacitectonite (*sensu* Benn & Evans 1996), possibly derived from LFA 2.

Log B. The coarsening upward gravel in Log B records highly turbid to hyperconcentrated discharge in non-channelized sheets. This is interpreted as aggrading glacifluvial outwash from an approaching ice margin. The capping sequence of crudely stratified diamicton and laminated sediments records a change from glacifluvial to subaqueous sedimentation prior to overriding by a glacier of unknown age that tilted the upper beds. The coarsening-upward sequence, apparent tectonism and capping diamict at Log B, clearly distinguishes these deposits from those of LFA 5a (west of Mary Sachs Creek, easternmost cliff, Fig. 3). Furthermore, although there is lithological similarity between Log B and LFA 5 throughout DHB, the sense of shearing is dissimilar (LFA 5 from the NE and Log B from the SE). The simplest explanation for this difference in shear direction is that LFA 5 (DHB) was displaced by interior ice advancing down the Kellett River whereas Log B was displaced by ice advancing along Amundsen Gulf (Fig. 2b).

Discussion

Large scale glacitectonic structures and stratigraphic architecture

The glacitectonic structures identified within the lithofacies associations at DHB record two phases of deformation (early N-S and later NE-SW). This deformation, together with the DHB stratigraphic architecture (Fig. 16), helps to explain both the kinetostratigraphy (*sensu* Berthelsen 1978) and the evolution of the local glacial geomorphology. The large scale glacitectonic deformation is most clearly manifest in the thrust faulting and conjugate shear development in the Kanguk Fm. bedrock and overlying LFA 1 (east cliff, Fig. 7d iv-vi). Here, predominantly northeasterly dipping thrust slices have been stacked, truncated and overlain by LFA 6 during the second and strongest (NE-SW) deformation phase (Fig. 16, note bedrock slice within the imbricated, stacked sequence of LFA 1). These structures demonstrate significant compression and thickening of the stratigraphic sequence, including ~40 m of vertical offset within the Kanguk Formation across ~3 km (Fig. 16). Prominent hydrofracture fills in LFA's 1 and 2 record elevated porewater pressure during this thrusting. Smaller scale shear zones have also developed in the finer grained interbeds within thrust slabs (Fig. 7d vii), indicative of deformation partitioning by strata prone to ductile failure. This zone of greatest displacement also coincides with the highest topography (Fig. 16). Lower angle thrust faults propagate in a southwesterly direction, predominantly through LFA 1 but are also developed through LFA 2 in the footwall block (east cliff, Fig. 16).

The thrust stacking of bedrock and LFA's 1-5 in east cliff (oblique to Fig. 16) represents the ice-proximal zone of a thrust-block moraine/composite ridge constructed by a glacier advancing from the northeast (i.e., the interior of Banks Island). This glacitectonic deformation propagated southwestward to distal parts of the proglacial stress field, as recorded by well-developed shear zones in the finer grained sediments of LFA's 2-5. In the more distal west cliff, open folds (LFA 2, Fig. 16) appear to be the product of a blind thrust within underlying LFA 1. This style of tectonism is typical of 'piggyback-style thrusting', wherein steeply inclined thrust blocks at the advancing ice margin propagate new thrusts in the footwall (Fig. 16; Park 1983; Mulugeta & Koyi 1987). The angle of these footwall thrusts becomes progressively shallower southwestward (ice-distal) in LFA 1-5, where deformation partitioning developed in finer grained lithofacies (Fig. 16).

The densely spaced ponds and depressions immediately to the northeast of DHB (Fig. 2b, 3) constitute the depression from which the thrust mass was excavated (e.g., Aber et al. 1989; Evans & England 1991; van der Wateren 1995). This interpretation is consistent with the strong northeast to southwest shear direction in the second phase glacitectonic structures and in LFA 6 clast macrofabrics . This indicates that construction of DHB occurred at the margin of a glacier advancing from the northeast (i.e., from Kellett River valley), overtopping DHB and depositing the LFA 6 glacitectonite. The continued advance of Kellett ice coalesced with a glacier in Amundsen Gulf marked by the interlobate Sachs Moraine containing ice-transported shells as young as 24 cal ka BP (machine age, 21,020±70 yrs BP, UCIAMS-89674), assigning this glaciation to the Late Wisconsin (Vaughan 2014). East of Mary Sachs Creek, the deformation of Log B outwash (from the southeast, Fig. 15, 16) likely records overriding by the same Amundsen Gulf glacier (coeval with LFA 6, Late Wisconsinan). During ice retreat, when Kellett River and Amundsen Gulf ice decoupled, DHB was exposed, allowing the northwesterly-flowing, proglacial outwash (LFA 5a) west of Mary Sachs Creek to extend to the shoreline marking postglacial marine limit at Kellett Point (11 m asl, Vaughan, 2014, Fig. 3). Collectively, this reconstruction supports a Late Wisconsinan age for the construction of DHB thrust block moraine, although its stratigraphy includes a range of ages. A palaeogeographic reconstruction of events based upon our interpretation of the sedimentology, stratigraphic architecture, tectonism and geomorphology is presented in Figure 17.

Comparison and revision of previous DHB model

DHB was previously interpreted to record horizontally bedded, Neogene to Quaternary sediments assigned to preglacial fluvial, and multiple glacial and interglacial episodes (Vincent 1982; 1983; 1990; Vincent et al. 1983; 1984; Barendregt et al. 1998). A composite palaeomagnetic record was also presented for this proposed stratigraphy, placing most of this deposition before the Brunhes/Matuyama boundary (Fig. 1c). However, prior to our work, detailed studies had not been undertaken on the sedimentology, structural geology or stratigraphic architecture of DHB. Rather, the original stratigraphy was restricted to a series of composite logs (our Fig. 4; reproduced from Vincent et al. 1983) that notably omitted the pervasive deformation structures including the thrust bedrock in east cliff (Fig. 16).

Our reinterpretation dismisses several key elements of the original DHB model, outlined below (oldest to youngest). We have determined that the lowermost unit (LFA 1) is not preglacial (Fyles 1990), but rather is composed of glacial outwash providing the earliest terrestrial evidence for glaciation in the area and containing allochthonous organics (including tree stumps) of unknown provenance. Because LFA 1 is a glacifluvial braidplain (Fig. 17a) it is not part of the Beaufort Fm. (Fyles 1990) even though it contains reworked macrofossils presumably derived from it (e.g., Craig and Fyles 1960; Hills et al. 1974; Vincent et al. 1983).

Our analysis of LFA 2 indicates that it is also not preglacial, nor can it be the Worth Point Fm. as previously proposed (cf. Vincent 1980, 1982, 1983, 1990; Vincent et al. 1983; Mathews et al. 1986). Indeed, the type locality of the Worth Point Fm. has been invalidated because it is now recognized to be an ice-transported and glacitectonized raft (Vaughan et al. this volume). In contrast, our LFA 2 is characterized by an isochronous tundra surface with ice wedges and intervening shallow ponds and wet meadows. The peat at LFA 2 is dominated by bryophytes (>95% by volume) and yet none of these macrofossils had been previously documented. The diverse, *in situ* bryophyte species record a flora similar to that on Banks Island today, suggesting that LFA 2 warrants an interglacial rank. This conclusion is supported by the limited vascular plant record (five taxa, no species indicated, Vincent 1990) that also suggest a wet meadow environment. The absolute age of LFA 2 remains unknown but is apparently older than the Brunhes/Matayama boundary (>780 ka; Fig. 18; cf. Vincent et al. 1984; Vincent & Barendregt 1990; Barendregt et al. 1998; see below).

LFA 3 in both our model and that of Vincent et al. (1983) provides the first direct evidence of ice reaching DHB. In the previous model this was interpreted as Bernard Till (Banks Glaciation) that was bracketed by glacioisostatically induced marine transgressions (pre- and post-Banks seas; Vincent 1982, 1983). Within basal LFA 3, we recognized highly deformed stratified sediment that was deposited in a subaqueous environment (deep water, Fig. 17c) prior to the arrival of overriding ice from the south (recorded by granite and gabbro erratics; Figs. 17c, 18). We regard LFA 3 to be equivalent to the Banks Glaciation (Vincent 1980, 1982, 1983; Vincent et al. 1983), including its pre- and post-Banks seas represented by the stratified sediments that bracket the central diamicton. The LFA 3 sediments are also reported to be magnetically reversed, thereby predating the Brunhes-Matayama boundary (>780 ka; Fig. 18; Vincent et al. 1984; Vincent & Barendregt 1990; Barendregt et al. 1998).

Following the Banks Glaciation (our LFA 3 at DHB), Vincent (1980, 1982, 1983) and Vincent et al. (1983) recognized two interglacials (Morgan Bluffs and Cape Collinson) separated by an intervening marine episode. This marine episode corresponded with the glacioisostatic loading of a Laurentide ice margin confined to the eastern half of Banks Island (Thomsen Glaciation). However, LFA 4 contains coarsening-upward subaqueous sedimentation from the south or southwest reflecting continued marine regression following LFA 3 (Fig. 17c, d). The coarsening-upward sequence of LFA 4 is supplanted by coarse gravel outwash (LFA 5) supplied by a glacier readvancing from the northeast (Kellett River valley ice; Fig. 17e). Glacier overriding is recorded by LFA 6 glacitectonite that caps the DHB thrust-block moraine composed of the entire assemblage of underlying LFA's (Fig. 16).

Notably, we have not observed any sediments characteristic of two interglacial deposits above LFA 3, hence dismiss the presence of the Morgan Bluffs and Cape Collinson interglacials at DHB. Nor have we observed intervening marine sediments attributable to the Thomsen Glaciation (e.g., 'Big Sea'; Vincent 1982, 1983; Vincent et al. 1983). Rather, following the deposition of the subaqueous sediments of upper

LFA’s 3 and 4 (post-Banks Glaciation), LFA’s 5 and 6 record a single ice advance reaching DHB from the Kellett River valley (Banks Island interior) that was not reported by Vincent et al (1983). Rather, it was assumed that after the Banks Glaciation, DHB lay distal to any later ice advances. The nearest subsequent Laurentide margin to DHB, recognized by Vincent et al. (1983), was assigned to the Sachs Moraine (immediately east of DHB) that was presumed to be of Early Wisconsin age (Fig. 2b). However, the Sachs Moraine is now dated to ≤ 24 cal ka BP (Late Wisconsinan; Vaughan 2014) and was formed by the coalescence of Amundsen Gulf ice and the Kellett River ice that deformed DHB (Fig. 17). On the east side of Mary Sachs Creek, the deposition of the Sachs Moraine coincides with the deformation (from the southeast) of the underlying Log B outwash gravel (Figs. 3, 15b).

Although interglacial deposits were previously reported from the cliffs east of Mary Sachs Creek (i.e., Cape Collinson Interglacial, Vincent et al. 1983) we found no similar evidence. Rather, the only deposit that we observed that may have been construed as interglacial sediments, occurs in our Log A (Figs. 3, 15A), Here, a dark, discontinuous band of allochthonous organics (Fig. 15A) sits severed within a heavily tectonized melange (type B glacitectonite) and is therefore clearly a raft. Although the absolute ages of the deformed units east of Mary Sachs Creek (Logs A & B) remain uncertain, they are apparently normally magnetized (Vincent et al., 1984; Barendregt & Vincent 1990). Blake (1987) also reports non-finite radiocarbon dates (>36,000 & 49,000 BP, GSC-3560 & 3560-2) obtained on “compressed and deformed black woody peat” purportedly from Cape Collinson beds (Vincent et al. 1983, p. 1708). Additionally, Causse & Vincent (1989) report a U-series date on shells of 92.4 ka (UQT-143) from beds similar to the Log B gravel. Again, all of these dated units have been displaced from unknown locations, limiting their utility for paleoenvironmental reconstruction. The last glacially influenced sedimentation at DHB is recorded by northwesterly directed outwash (easternmost cliff, LFA 5a; Figs. 3, 16, 17g). We infer that this outwash fed the gravel beach marking marine limit (11 m asl) in the lower Kellett River during Late Wisconsinan retreat from the Sachs Moraine.

Regional context

The stratigraphy at Duck Hawk Bluffs records three glacial events. The earliest of these is recorded by the aggradation of a glacifluvial braidplain (LFA 1) that contains allochthonous preglacial material of diverse age. However, within LFA 1 there is no evidence that the ice reached DHB, although it must have been nearby (southeast, in Amundsen Gulf). The age of this glaciation remains unknown but the paleomagnetic polarity of LFA 1 indicates that it is >780 ka BP. (Barendregt & Vincent 1990). Although Barendreght et al. (1998) assigned part of the Worth Point Fm to the normally polarized Olduvai sub-zone (MIS 64-74), it should be noted that the Worth Point section is located 30 km northwest of DHB. There, a stratigraphic revision has demonstrated that this "type section" of the Worth Point Fm is in fact a glacitectonized raft (Vaughan et al., this volume). This observation invalidates the previous correlation from Worth Point to purportedly coeval beds at DHB (our LFA 2). Furthermore, the diverse plant assemblages in the Worth Point raft include preglacial tree stumps and *Sphagnum* spp. that do not occur on Banks Island today and are not found in the wet meadow assemblages (rich fen) of LFA 2 that we interpret to be of interglacial rank. Therefore, the sediments at DHB (LFAs 1 & 2) do not support an age assignment of MIS 64-74 (Olduvai subzone) proposed by Barendregt et al. (1998). Currently, the available paleomagnetic record at DHB simply separates the entire section into a lower 'reversed' unit and an upper 'normal' one. We place this boundary within our LFA 4 (Fig. 18).

The overriding of DHB by glaciers is recorded in LFAs 3 & 6. These two glacial deposits may correspond to parts of the record of stacked till sheets reported from the offshore seismic stratigraphy (Batchelor et al. 2012). LFA 3 glaciation may have contributed to the onset of mid-Pleistocene ice-rafted debris (IRD, Ca-rich) attributed to the Canadian Arctic Archipelago (CAA) and recorded in sediment cores from the Arctic Ocean basin (Stein et al., 2010; Polyak et al., 2009, 2013; O'Regan et al., 2010). However, Stein et al. (2010), place the onset of Ca-rich ice-rafted debris within MIS 16 (~659 ka BP; Lisiecki & Raymo, 2005) with subsequent peaks at MIS 12, 10 and 8, all of which postdate the >780 ka BP paleomagnetic age proposed for LFA 3 (above). Polyak et al. (2013) place the onset of Ca-rich ice-rafted debris in the central Arctic Ocean closer to MIS 19 (790 ka BP), which is broadly compatible with the DHB paleomagnetic record. LFA 6 records the last glaciation of DHB, and we assign this to the Late Wisconsinan, when Banks Island ice (Kellett River) coalesced with the Amundsen Gulf Ice Stream (Stokes et al. 2006; MacLean et al., 2013). During this interval, the northwest Laurentide Ice Sheet likely reached the edge of the polar continental shelf (Stokes et al, 2005, 2006; England et al., 2009; Batchelor et al., 2012; MacLean et al., 2013).

Conclusions

New sedimentological and stratigraphic analyses of DHB fundamentally revise the previous reconstruction of the Neogene and Quaternary history of Banks Island. DHB, and nearby sections (Vaughan et al. this volume), have been widely cited as important, undeformed terrestrial archives of glacial and interglacial sedimentation at the northwest limit of the Laurentide Ice Sheet. The recognition of pervasive, large-scale deformation of DHB necessitates a fundamental re-assessment of its previously described layer-cake stratigraphy. We document that the oldest sediments (LFA 1) at DHB are not preglacial fluvial deposits (Beaufort Fm) but rather record proglacial outwash from ice in Amundsen Gulf that did not reach the site. This was followed by the establishment of an isochronous tundra surface of interglacial rank (LFA 2), characterized by ice wedge polygons within a wet meadow that supported a bryophyte community similar to modern. Subsequently, glacial sedimentation resumed, initially in an ice-proximal, subaqueous environment followed by ice arrival and glacitectonic deformation (LFA 3). The overlying deposits at DHB indicate post-glacial, marine sedimentation (LFA 4), followed by the aggradation of glacifluvial sand and gravel (LFA 5) from an advancing ice margin that culminated with the emplacement of a glacitectonite (LFA 6) during the Late Wisconsin. This ice advance constructed a previously undescribed thrust-block moraine (60 m high, 8 km long) that exhibits pervasive deformation, and includes a substantial raft of Cretaceous Kanguk Fm bedrock (Fig. 16). Furthermore, the production of the glacitectonite (LFA 6) cannibalized LFA's 1 to 5 which at some sites were previously thought to be *in situ* interglacial deposits. Finally, outwash (LFA 5a) incised the DHB thrust block during deglaciation of the site when ice in the Kellett River and Amundsen Gulf separated and proglacial meltwater drained northwestward to marine limit (11 m asl). Refinements to the evolution and chronology of DHB, and other sections around Banks island, will further contribute to the understanding of high latitude environmental change, especially the complementary marine archives of the adjacent polar continental shelf and Arctic Ocean.

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Figure captions

Figure 1: Banks Island and the traditional Quaternary stratigraphy: a) location map of Banks Island in the western Canadian Arctic; b) map of the traditional proposals for the extent of glaciations based on surficial geology and Quaternary stratigraphy (from Vincent 1983). WPB = Worth Point bluff, DHB = Duck Hawk Bluffs, VMSIS = Viscount Melville Sound Ice Shelf moraine location, MB = Morgan Bluffs, NRB = Nelson River bluff; c) summary of traditional stratigraphy, magnetostratigraphy and reconstructed glacial and interglacial events based on the Duck Hawk Bluffs Formation from SW Banks Island (after Barendregt & Vincent 1990). Note that this is a composite section based upon Vincent et als. (1983) logs A-I in the "west" and "central" cliffs and that the upper details labelled "Bluffs E Mary Sachs Ck" are from their log J in the "Mary Sachs Creek Cliff" (see Figures 3 & 4).

Figure 2: Aerial views of the Duck Hawk Bluffs: a) oblique aerial photograph of the bluffs viewed from the east; b) vertical aerial photograph extract of the local terrain of the southwest coast of Banks Island (photograph A17381-41, National Air Photograph Library, Ottawa). Black broken line arrows are relict meltwater channels. Areas outlined by red dotted lines are likely glacitectonic thrust masses based on interpretations of the sedimentology and stratigraphy reported in this paper. The "Sachs Till" moraine of Vincent (1983) is outlined by white dotted line.

Figure 3: Topographic map of the SW coast of Banks Island showing the locations of the cliff exposures examined during this study. Each cliff contains one or more vertical profile log location, which are designated by letters (see Figure 7).

Figure 4: Vincent et als. (1983) lithostratigraphy for the Duck Hawk Bluffs (logs A-I) and Mary Sachs Creek cliff (log J). Logs A-G are in our West cliff and logs H and I are in our Central cliff.

Figure 5: Clast form co-variance plots of data from this study: a) Type 2 clast form co-variance plot from

Lukas et al. (2013), used in this study as control sample data for the interpretation of clast wear patterns and former transport histories; b) C40/RA and C40/RWR co-variance plots for data collected in this study.

Figure 6: Quantitative plots of clast fabric strength: a) clast fabric shape ternary plot (Benn, 1994) containing A/B plane (blue) and A axis (red) data used in this paper and control sample data (A/B planes) for glacitectonite (Evans et al.,1998; Hiemstra et al., 2007), subglacial till (Evans and Hiemstra, 2005) and lodged clasts (Evans and Hiemstra, 2005); b) modality-isotropy plot (Hicock et al.,1996; Evans et al., 2007) of the clast A/B plane (blue) and A axis (red) macrofabric data used in this paper. Envelopes contain data from deposits of known origin and shaded area represents that part of the graph in which stronger modality and isotropy in subglacial traction tills or glacitectonites reflects an increasing lodgement component (based on Evans and Hiemstra, 2005; Evans et al., 2007). This graph is thereby used to interpret trends in cumulative strain signature in the glacitectonite-subglacial traction till continuum (un, unimodal; su, spread unimodal; bi, bi-modal; sb, spread bi-modal; mm, multi-modal). See Figure 7 for sample locations.

Figure 7a: Stratigraphic logs and sedimentological details of westernmost cliff: i) Log A; ii) Log B iii) photomosaic of lower Log B; iv) photomosaic of upper log B; v) Log C; vi) Log D. In all logs the single barbed arrows with compass orientations indicate sense of shearing based on thrust planes and the solid arrows with compass orientations indicate palaeocurrent directions.

Figure 7b: Stratigraphic logs and sedimentological details of west cliff: i) Log A1; ii) Log A; iii) Log B; iv) Log C; v) Log D.

Figure 7c: Stratigraphic logs and sedimentological details of central cliff: i) Log A; ii) details of deformation structures in LFA 6 at top of Log A; iii) supplementary details of upper part of sequence located immediately east of Log A.

Figure 7d: Stratigraphic logs and sedimentological details of east cliff: i) Log A; ii) photomosaic of lower Log A showing lower LFA 1 details; iii) photomosaic of middle Log A showing upper LFA 1 details (inset photos l & m) and LFA 2 details (inset photos a-k); iv) overview of the eastern half of east cliff, showing locations of logs A and B, a photolog of the capping gravelly glacitectonite carapace, major faults and the LFA 1/Kanguk Formation contact. The anomalous dips of the LFA 1 bedding associated with the major faults are represented by the 3-D plane symbol ; v) overview of the deformation structures in the eastern part of the east cliff, showing prominent distorted bedding, major thrusts and 3-D representations of anomalous dips in LFA 1 bedding. The approximate boundary between the Kanguk Formation bedrock and LFA 1 provides an outline of a prominent thrust block that has been displaced along a major fault descending below beach level at the far right of the image; vi) details of the major fault structures and chevron folding developed in the thrust block of Kanguk Formation, with the shearing direction represented by the thrust plane dip orientations in a lower hemisphere stereoplot; vii) Log B compiled as a photomosaic of deformation structures developed within sand/silt interbeds in the LFA 1 strata that overlie the Kanguk Formation thrust block; viii) details of deformation structures identified in the Kanguk Formation thrust block.

Figure 7e: Stratigraphic logs and sedimentological details of easternmost cliff: i) photomosaic of the easternmost cliffs, showing locations of the three vertical logs; ii) Log A; iii) Log B; iv) Log C, showing details of cobble clast lag at contact of LFAs 1 and 5 in inset “c” and a vertical, branching clastic dyke in LFA 1 in inset “d”.

Figure 8: Details of LFA 1: a) example of a concentration of logs with degraded rootballs in the stacked tabular units of cross-bedded sands and gravels and crudely stratified matrix-supported cobble to boulder gravels in the west cliff; b) poorly-sorted and crudely horizontally bedded, cobble to boulder gravel, containing laminated clay intraclast (immediately above compass); c) shallow,

locally graded foreset beds composed of openwork to sandy matrix-supported cobble to granule gravels in central cliff Log A; d) planar bedded sandy granule gravels with gravel lags (top and bottom of image) interrupted by an intervening unit of cross-bedded, poorly-sorted to openwork cobble to boulder gravel, crudely planar-bedded granule to pebble gravel and massive cobble to pebble gravel. A cluster of small tree fragments occur at the base of the cobble to boulder gravels at image centre; e) middle sequence of stacked tabular units of predominantly horizontally bedded sands and gravels in DHB west cliff Log D; f) sequence of fining-upward, planar-bedded cobble to sandy, fine gravels, containing clusters of wood detritus in the coarser gravel beds (central cliff Log A); g) planar cross-bedded, sandy granule to pebble gravels with pebble to cobble lags and wood fragments (central cliff Log A).

Figure 9: Details of LFA 2 from the west cliff: a) narrow necked ice wedge pseudomorph with upturned marginal bedding in sand, silt and clay rhythmites; b) wide necked ice wedge pseudomorph; c) ice wedge pseudomorph with overturned silt, sand and clay rhythmites, developed at the top of LFA 2 and sealed by overlying LFA 3 laminated diamicton; d) horizontally bedded sand and silt containing well preserved moss macrofossils; e) zone of disharmonically folded sand/silt laminae with well preserved moss peat, located at the base of LFA 2.

Figure 10: Details of simple shear structures developed in fine sand and silt laminae at the centre of the thickest outcrop of LFA 2 in west cliff Log B: a) overview of large scale anastomosing shear faults with zones of smaller scale, densely spaced anastomosing shears towards the base of the photograph; b) details of large scale shear faults separating zones of variably but predominantly densely spaced shears; c) close details of shear zone comprising densely spaced anastomosing fractures; d) details of ascending kink zone cross-cutting large scale anastomising shear faults. Shearing direction is coming out of the cliff and towards the right in each image.

Figure 11: Details of LFA 2 near west Log D. Lower box shows a 1-2m thick basal zone comprising highly

attenuated inter-digitated beds of gravelly to clay-rich diamictons and sand/silt/clay rhythmites. Upper box shows rhythmites containing diamictic intraclasts with sharp and angular boundaries; also visible is the draping and deforming of the rhythmite bedding over and under the intraclasts respectively. “Tails” extending from the diamictic intraclasts and thin but discontinuous diamictic beds are also visible in the upper box.

Figure 12: Details of LFA 4 in West Log A: a) heavily deformed upper contact of LFA 4 comprising sand/silt/clay rhythmites, displaying well developed boudinage, sheath folds and immature tectonic laminae and an interdigitated/sheared boundary with underlying climbing ripple sands. The contact with overlying LFA 6 is marked by an interdigitated zone containing rooted and de-rooted folds or rhythmite rafts in a sand, gravel and diamicton mélange; b) narrow, anabranched clastic dykes ascending sub-vertically through climbing ripple drift and resulting in offset bedding between blocks of host material; c) abrupt contact between sheared sands and rhythmites of LFA 4 and overlying sand, gravel and diamicton mélange.

Figure 13: Details of LFA 5: a) poorly sorted and matrix-supported, pebble to cobble gravel; b) tabular sequence of horizontally bedded to massive gravel and matrix-supported gravel containing a horizon of striated cobbles and boulders; c) heavily deformed, discontinuous bed of silt/clay rhythmites between LFA 5 and LFA 6 in West Log D, showing the amalgamation zone with the LFA 5 gravels immediately above the compass; d) interdigitated/deformed contact between LFA 5 granule to pebble gravels and underlying LFA 4 laminated silts and clays with organic detritus in West Log B.

Figure 14: Details of the typical moss peat and associated microfossil materials at Duck Hawk Bluffs: a) images of *Calliergon richardsonii*, the most abundant bryophyte in LFA 2; b) modern southern Banks Island analogue of a rich tundra fen environment for the bryophyte assemblages of LFA 2; c) degraded organics typical of the materials from LFA 6.

Figure 15: The stratigraphy of Mary Sachs Creek cliff: a) photographic compilation log A, showing upward-fining, highly contorted, cross-stratified sands and fine gravels to rhythmically bedded sands, silts, clays and minor pebble gravels. Also visible is a lens of interbedded laminated silty clays with dropstones and gravelly diamictons containing sandy, black coloured organic material with wood fragments. An ice wedge pseudomorph is visible to the right of the exposure; b) photographic compilation log B showing: i) laminated to crudely stratified, cobbly but silt/clay-rich diamicton, grading into interlaminated silts, clays and pebbly silty sands; ii) poorly-sorted cobble to boulder gravel and bouldery matrix-supported gravel; iii) coarsening-upward sequence of horizontally bedded sandy granule gravels with occasional silty sand laminae; iv & v) pebble to cobble gravel and matrix-supported gravel.

Figure 16: Stratigraphic cross-profile, running west to east, of Duck Hawk Bluffs based upon interpolations between the main section logs and showing the six LFAs, the Kanguk Formation bedrock exposures, major structural features and the positions of ice wedge pseudomorphs and significant clastic dykes. At the eastern end of the cross-profile, the sediments at the core of Mary Sachs Creek cliff log A are tentatively classified as LFA 2, although these materials are likely not in situ. Note that the cross-profile does not extend as far east as Mary Sachs Creek cliff log B.

Figure 17: The palaeogeography of southwest Banks Island based upon interpretations of the principal lithofacies associations and structural architecture recognized at Duck Hawk Bluffs and the geomorphology of the surrounding terrain. See text for detailed explanations.

Figure 18: A revised lithostratigraphy for Duck Hawk Bluffs. Previously reported age constraints are also depicted together with a tentative allocation of MIS stages (positioned alongside East Cliff Log A for clarity) based upon the palaeomagnetic record.